Contents lists available at ScienceDirect



Earth and Planetary Science Letters



journal homepage: www.elsevier.com/locate/epsl

Competition between erosion and reaction kinetics in controlling silicate-weathering rates

G.E. Hilley ^{a,*}, C.P. Chamberlain ^b, S. Moon ^a, S. Porder ^c, S.D. Willett ^d

^a Department of Geological and Environmental Sciences, Stanford University, Stanford, California 94305-2115, USA

^b Department of Environmental Earth System Sciences, Stanford University, Stanford, California 94305-2115, USA

^c Department of Evolutionary Biology and Ecology, Brown University, Providence, RI 02912, USA

^d Geological Institute, Eldgenössische Technische Hochschule Zürich, 8092 Zürich, Switzerland

ARTICLE INFO

Article history: Received 27 March 2009 Received in revised form 4 January 2010 Accepted 6 January 2010 Available online 16 March 2010

Editor: M.L. Delaney

Keywords: chemical weathering landscape evolution silicate mineral weathering geologic carbon cycle

ABSTRACT

Weathering rates of minerals may be limited by the reaction kinetics of the constitutive minerals or the supply of fresh minerals delivered to the near surface by denudation. We use a model of denudation and reaction kinetics to analyze the relative importance of each in moderating silicate-weathering fluxes in different erosional environments for different, commonly occurring silicate minerals. We find that minerals that reside in the near-surface weathering zone for far longer than the reaction takes to complete, as is the case with slow denudation, produce weathering fluxes that respond most strongly to changes in denudation rate. Conversely, when transit through the weathering zone is rapid relative to a given mineral's weathering timescale, the weathering responds most vigorously to changes in the reaction kinetics. Based on field- and laboratory-measured values of silicate-weathering kinetic constants, observed weathering zone thicknesses in relation to denudation rates, and denudation rates inferred from topography, it appears that fresh mineral supply may play a sub-equal or dominant role in moderating silicate-weathering fluxes for four mineral phases considered. If correct, this suggests that the concentration of atmospheric CO₂, which is regulated by silicate weathering over geologic timescales, may depend on those factors that control long-term erosion rates across Earth's surface.

© 2010 Elsevier B.V. All rights reserved.

1. Introduction

Weathering rates of minerals may be limited by the reaction kinetics of the constitutive minerals or the supply of fresh minerals delivered to the near surface by denudation. Should reaction kinetics limit weathering rates, changes in environmental factors such as temperature should likewise alter the reaction rates of these minerals. However, if minerals are in short supply because reactions have proceeded to near completion, tectonic processes, which generally produce accelerated denudation rates, may enhance the supply, and thus net weathering rates of these minerals (Berner et al., 1983; Stallard and Edmond, 1983; Raymo et al., 1988; Berner, 1990; Raymo and Ruddiman, 1992; Velbel, 1993; Berner, 1994; Brady and Carroll, 1994; Lasaga et al., 1994; White, 1995; White and Blum, 1995; White and Brantley, 1995; Ruddiman et al., 1997; Lasaga, 1998; Riebe et al., 2003; Chamberlain et al., 2005; Waldbauer and Chamberlain, 2005; Ferrier and Kirchner, 2008; Gabet and Mudd, 2009). Mineral phases react at different rates, and so depending on the mineral reaction time versus the residence time of rock in the weathering zone, different mineral phases may be limited by each of these effects in a particular situation (Brantley, 2008). In this contribution, we use a simple modeling approach with a compilation of our current state-of-knowledge about mineral reaction kinetics, denudation rates, and the vertical extent of weathering in the subsurface to provide a coarse assessment of the relative importance of reaction kinetics and denudation in regulating chemical fluxes produced by mineral supply and weathering for different silicate mineral phases.

2. The model

Because of the large uncertainties in both the nature of chemical/ denudation processes and their rates, we have opted to use a model that treats these processes in the simplest way possible, while still retaining some coarse realism of the chemistry and geomorphology that may be applicable at regional scales. The core of the model has been developed and presented elsewhere (Chamberlain et al., 2005; Ferrier and Kirchner, 2008; Gabet and Mudd, 2009; Hilley and Porder, 2008; Hren et al., 2007b; Waldbauer and Chamberlain, 2005); herein, we use the model to estimate the relative importance of denudation and reaction rates, and the scaling of the relevant parameters.

We consider that weathering occurs within a surface layer of thickness Z (Fig. 1). Material enters this layer at the base and exits at

^{*} Corresponding author. E-mail address: hilley@stanford.edu (G.E. Hilley).

⁰⁰¹²⁻⁸²¹X/\$ - see front matter © 2010 Elsevier B.V. All rights reserved. doi:10.1016/j.epsl.2010.01.008



Fig. 1. Schematic model depiction showing how denudation and chemical weathering processes moderate mineral weathering rates.

the surface by denudation at a rate of ε [L t⁻¹]. While in transit through this layer, reactant is depleted at a rate that depends on a kinetic rate constant, k_i [mol L⁻² t⁻¹], a mineral surface area, A_i [L² mol⁻¹], and the concentration of the reactant phase at a particular time $[mol L^{-3}]$. Likewise, the rate at which weathering products are generated scale with the reactant loss rate, multiplied by a stoichiometric ratio, r_i . Source material for the *i*th reaction has an initial concentration prior to weathering of $_{o}q_{i}$ [mol L⁻³]. The flux of each mineral phase that is formed by the weathering reaction, $_pF_i$ $[mol L^{-2} t^{-1}]$ is the product of the denudation rate and the concentration of the product species at the surface (z=Z) (Chamberlain et al., 2005; Waldbauer and Chamberlain, 2005; Hren et al., 2007b; Hilley and Porder, 2008; Gabet and Mudd, 2009), where Z is the thickness of the weathering zone. The timescale of the reaction kinetics is defined by $(k_iA_i)^{-1}$, which has been inferred independently from field and laboratory studies (Brantley et al., 1986, 1993; White, 1995; White and Blum, 1995; White and Brantley, 1995; White et al., 1996, 2001; White and Brantley, 2003). The specific values for this timescale depend on the mineral phase of interest (White and Brantley, 1995; White et al., 1996), the timescale over which weathering occurs (White and Brantley, 2003), and the flow-through of water (Clow and Drever, 1996; Maher et al., 2007). For simplicity, we assume that for a given mineral phase undergoing a particular hydrologic regime, $k_i A_i$ can be represented by a particular, empirically specified value that does not change over time. We acknowledge that this is a strict assumption indeed: a comparison of laboratory and field kinetic rates shows that this product likely decreases over time due to a variety of processes, including mineral surface chemistry (Brantley et al., 1986, 1993), hydrology (Clow and Drever, 1996), and saturation state (Lasaga et al., 1994; Lasaga, 1998). In addition, biological processes may reduce the pH within pore waters and soil, and this may alter the bulk reaction kinetics in natural systems. Thus, a myriad of physical, chemical, and biological processes control the values of k_i and A_i , which we simply subsume into a constant, specified value of *k_iA_i* for a particular mineral phase. In the analysis below, we consider a broad range of these values that spans both field- and laboratorybased measures of these kinetic rate constants in an attempt to capture the natural variations in weathering kinetics that arise from these complexities (see below; Table 1).

Given the simplifying assumption of unchanging k_iA_i over time, the concentration of the product species at the surface is found by solving the advection–reaction problem for material traversing the weathering layer. This is best done by considering the non-dimensional problem, obtained by scaling length by *Z* and time by Z/ε , which is the time over which rock resides in the weathering zone. While ε may be

Table 1

Range of laboratory- and field-based reaction rates used in this study (compiled from White and Brantley, 2003).

Mineral phase	Laboratory-based	Field-based		
	$log_{10} k_i A_i$ range (1/s)	$\log_{10} k_i A_i$ range (1/s)		
Albite K-feldspar Hornblende Biotite	- 10.9 to - 9.2 - 11.4 to - 9.0 - 10.6 to - 8.3 - 10.1 to - 8.1	-13.3 to -10.9 -14.4 to -13.4 -14.2 to -12.9 -13.2 to -11.0		

estimated using methods such as cosmogenic radionuclide analyses (Riebe et al., 2001a,b, 2003, 2004a,b), as discussed below, data on *Z* is far scarcer. Using these quantities, the non-dimensional product flux is defined as:

$${}_{p}F_{i}^{*} = \frac{{}_{p}F_{i}}{{}_{o}q_{i}r_{i}\varepsilon}$$
(1)

which expresses the product flux from the surface normalized by the flux of reactant into the base of the weathering layer (adjusted for the mineral phase's stoichiometric ratio), so that a value of ${}_{p}F_{i}^{*} = 1$ implies that all source material has been weathered. In steady-state, the non-dimensional flux of the *i*th reaction product from the surface of the eroding layer is given as:

$${}_{p}F_{i}^{*} = 1 - \exp(-D_{i}) \tag{2}$$

where D_i is the Damköhler number for the *i*th reaction and is defined as the dimensionless group:

$$D_i = \frac{k_i A_i Z}{\varepsilon} \tag{3}$$

The Damköhler number for a particular mineral reaction can also be interpreted as the ratio of the residence time of minerals in the weathering zone (Z/ε) to the reaction timescale $(k_i^{-1}A_i^{-1})$ (White and Brantley, 2003). The fact that this dimensionless group is the only parameter in Eq. (2) demonstrates that this is the fundamental scaling number relating mineral supply and reaction kinetics.

Much effort has been invested in estimating kinetic parameters, as discussed below, but the thickness of the weathering zone is also of first order importance and its estimation is problematic. We expect that Z varies with a number of factors, including the permeability of the rock (Clow and Drever, 1996; Porder et al., 2007), the loss of permeability through precipitation of secondary mineral phases (Maher et al., 2007), subsurface temperature, biological activity, and erosional processes (Raymo and Ruddiman, 1992; Ruddiman et al., 1997). The role that these factors play is poorly understood because of the difficulty in accessing the entire weathering zone and relating its downward propagation to specific hydrologic and environmental factors. However, as we argue below, there are probably physical relationships between the erosion rate and the weathering zone thickness.

In the simplest model, the weathering zone may deepen at a constant rate (Drever and Clow, 1995; White and Brantley, 2003) that depends on the local hydrology, chemistry, and biological processes present at a given site. However, this simple idealization leads to the uncomfortable conclusion that areas in which erosion rates exceed this downward propagation rate should quickly become devoid of weathered material, exposing fresh unweathered bedrock everywhere at the surface. Conversely, should erosion rate be less than this downward propagation rate, the weathering zone would thicken indefinitely, leading to the removal of all primary minerals at the surface over time as mineral residence times increase along with the weathering zone thickness. The fact that we observe weathering zones and mineral detritus that varies somewhat systematically, but

not dramatically, between actively eroding and stable areas suggests that there must be some negative feedback between weathering zone thickness and erosion rate. In the related case of development of a mobile soil layer, this feedback has been quantified in a rule of the form (Carson and Kirkby, 1972; Heimsath et al., 1997, 1999; Dietrich et al., 2003):

$$\frac{dh}{dt} = B_{o_{\text{soil}}} \exp(-B_{b_{\text{soil}}}h) \tag{4}$$

where *h* is the thickness of the mobile soil layer [L], $B_{o_{\text{soil}}}$ is the rate at which the soil thickens when bedrock is exposed at the surface [L/t] and $B_{b_{\text{soil}}}$ is a scale factor that describes dependence of the thickening rate to the soil thickness [L⁻¹]. Although the physically perturbed soil layer and the weathering zone are not equivalent, we assert that feedback mechanisms also regulate the weathering layer thickness and we can therefore model the weathering layer thickness with Eq. (4) (as suggested by Lebedeva et al., 2007), provided the appropriate parameters for the weathering zone, which we denote B_o and B_b , are determined in place of $B_{o_{\text{soil}}}$ and $B_{b_{\text{soil}}}$. In a landscape in which erosion removes material from the surface, the rate of change of the weathering layer thickness (*Z*), and adding the denudation rate to the right side this expression. Under these conditions there is a steady state $(dZ/dt = \varepsilon)$ in which the layer thickness is:

$$Z = -\frac{1}{B_b} \ln\left(\frac{\varepsilon}{B_o}\right) \tag{5}$$

If the weathering layer thickness is set by Eq. (5), the progression of the weathering reactions and the weathered product flux can be calculated as in Eqs. (1)-(3), but we can now express the Damköhler number in terms of the production coefficients in Eq. (5) as:

$$D_i = -\frac{k_i A_i}{B_b \varepsilon} \ln\left(\frac{\varepsilon}{B_o}\right) \tag{6}$$

This suggests the use of the ratio between denudation rate and the maximum weathering layer advance rate, which we denote as:

$$\varepsilon^* = \frac{\varepsilon}{B_o} \tag{7}$$

This allows us to express the flux of weathering product (Eq. (2)) as:

$${}_{p}F_{i}^{*} = 1 - \exp\left[\frac{A_{i}k_{i}}{B_{b}B_{o}\varepsilon^{*}}\ln(\varepsilon^{*})\right]$$
(8)

or,

$$\ln(1 - {}_pF_i^*) = \frac{A_i k_i}{B_b B_0 \varepsilon^*} \ln(\varepsilon^*)$$
(9)

Both the dimensionless weathering product flux ${}_{p}F_{i}^{*}$, and the ratio, ε^{*} , vary from 0 to 1 and Eq. (9) demonstrates that in fact these quantities directly co-vary with a coefficient, $\frac{A_{i}k_{i}}{B_{i}B_{o}}$.

A model based on Eq. (4) is largely empirical and application to development of the weathering layer depends on the similarity between soil-forming processes and downward propagation of the weathering zone front into bedrock. An alternative theory is based on reactive-transport modeling (Maher et al., 2007; Lebedeva et al., in press). If mass balance in the system is dominated by downward transport of reaction products by pore-water flow, the advance of a weathering front depends primarily on the magnitude of the water flux (Maher et al., 2007). This provides a parametric link between the vertical extent of weathering, hydrology, and geomorphic processes.

The dependence of weathering extent on denudation rates is not replaced by considering hydrologic processes; it is simply made more complex as there are clear links between denudation rates, climate, and the topographic relief that drives both erosional processes and subsurface fluid flow.

In this case, we note that the base level that is set by the major regional rivers largely limits the extent to which surface waters penetrate vertically into rock. For example, periodic variations in surface topography that allow enhanced vertical water flow along ridges in the landscape are rapidly damped with depth in favor of horizontal transport and draining of the landscape into channels (Turcotte and Schubert, 2002). Thus, we posit that the topographic relief sets the scale for the subsurface hydrologic system, which in turn determines the scale of the weathering zone. Denudation rates also correlate with topographic relief, and this correlation provides a link between erosion rate and the weathering depth scale. Following the work of Ahnert (1970) and Montgomery and Brandon (2002), denudation rate is correlated to relief as a power function:

$$\varepsilon = k_1 R_f^{k_2} \tag{10}$$

where R_f is the local relief at a characteristic scale, and k_1 and k_2 are empirically calibrated constants. If the relief in Eq. (10) characterizes the important topography that drives fluid flow in the weathering zone, we can thereby equate R_f with Z. In so doing, we are neglecting the permeability structure of the subsurface; in particular, we implicitly permit water to freely penetrate deeply into the basement as the relief can be on the order of kilometers. In many catchments, it is unlikely that the weathering zone extends to a depth equal to R_{f} , as precipitation is routed through a shallow conductive layer and ultimately is discharged into channels (Montgomery and Dietrich, 1988; Dietrich et al., 1992; Montgomery and Dietrich, 1992). However, in some high relief mountainous regions, there is some evidence for deeply penetrating fluids reacting with the host rock at depths of several kilometers (Evans et al., 2001). In well-drained, lowrelief landscapes, this model is likely to provide an upper limit depth scale to groundwater penetration. Given the supposition that the local relief sets an upper limit on weathering zone thickness, we rearrange Eq. (3) to recast D_i solely in terms of erosion rate:

$$D_i = \frac{k_i A_i}{k_1^{(1/k_2)}} \varepsilon^{(1/k_2 - 1)} \tag{11}$$

At low denudation rates, Montgomery and Brandon (2002) and Ahnert (1970) found that denudation and relief are linearly related to one another ($k_2 = 1$). In this case, D_i becomes independent of erosion rate:

$$D_i = \frac{k_i A_i}{k_1} \tag{12}$$

Thus, Eq. (12) indicates that changes in erosion rate are offset by changes in weathering zone thickness as the relief changes in the same direction. However, the overall weathering product flux retains a dependence on the denudation rate through the supply of new material to be weathered as is clear by noting the dimensional weathering product flux,

$${}_{p}F_{i} = {}_{o}q_{i}r_{i}\varepsilon \left[1 - \exp\left(-\frac{A_{i}k_{i}}{k_{1}}\right)\right]$$
(13)

Thus, even when the water balance plays a primary role in controlling the weathering zone thickness (Maher et al., 2007; Lebedeva et al., in press), we expect both reaction kinetics and denudation rate to play a role in determining the chemical weathering mass flux.

3. Results

3.1. General model behavior

We begin by considering the behavior of the model when ε and Z are allowed to vary independently. In this case, the non-dimensional flux can be cast in terms of the erosional Damköhler number described by Eq. (3). The graph of Eq. (2) is shown in Fig. 2A and demonstrates how the Damköhler number defines regions of parameter space where reactions have nearly gone to completion $({}_{p}F_{i}^{*}=1)$ and regions where little reaction has occurred $({}_{p}F_{i}^{*}=0)$. Large values of D_{i} produce ${}_{p}F_{i}^{*}=1$, while small values of D_{i} result in ${}_{p}F_{i}^{*}=0$.

Erosion rate appears in both D_i and in the normalized flux, ${}_{p}F_i^*$ (Eq. (1)), and so to explicitly illustrate the flux dependence on erosion rate, we remove ε from the non-dimensional flux through multiplication, and simply contour the quantity as a function of ε and $k_i A_i Z$ (Fig. 2B). In the case where ε greatly exceeds $k_i A_i Z_i$ the flux becomes insensitive to ε and instead varies in proportion to k_iA_iZ . Conversely, when $k_i A_i Z$ greatly exceeds ε , the flux depends solely on ε . These endmember conditions can be described algebraically by noting the limits as D_i ($k_i A_i Z/\epsilon$) goes to zero or infinity. In the case of the former, ${}_{P}F_i/\epsilon$ $(_{0}q_{i}r_{i}) = k_{i}A_{i}Z$ (White, 1995; West et al., 2005; Hren et al., 2007b), while in the case of the latter, ${}_{p}F_{i}/({}_{o}q_{i}r_{i}) = \varepsilon$ (West et al., 2005; Hren et al., 2007b). Thus, when rock resides far longer in the weathering zone than the reaction takes to proceed, the weathering flux is entirely dependent on the rate of mineral supply, a condition referred to as "supply-limited" (Fig. 2B). Alternatively, when rock resides within the weathering zone for a time much shorter than the kinetic timescale, the magnitude of the flux is purely a function of the reaction kinetics and layer thickness, referred to as "reaction-limited" conditions (Fig. 2B) (Stallard and Edmond, 1983; Waldbauer and Chamberlain, 2005; Hren et al., 2007b; Hilley and Porder, 2008; Gabet and Mudd, 2009). The role of reaction kinetics versus mineral supply will vary depending on the kinetics of a given mineral phase and so a given weathering zone may contain some minerals whose reaction rates are limited by fresh mineral supply, while other minerals may be limited by the mineral reaction rates.

In a second step, we show the dependence of ${}_{p}F_{i}^{*}$ on ε^{*} when the thickness of the weathering zone and denudation rate are logarithmically related to one another. In this case, the erosional Damköhler

number is represented by Eq. (6). $_{p}F_{i}^{*}$ decreases with increasing ε^{*} : the rapidity of this decrease is dictated by $k_{i}A_{i}/(B_{o}B_{b})$ (Fig. 3A).

As before, the flux term ${}_{p}F_{i}^{*}$ implicitly includes erosion rate, and so to consider the explicit dependence of erosion rate on flux, we can redimensionalize Eq. (9) using ε , B_{o} , and $k_{i}A_{i}/B_{b}$ and plot the flux as a function of the other dimensional parameters including erosion rate (Fig. 3B). When ε is small relative to B_{o} , ${}_{p}F_{i}/({}_{o}q_{i}r_{i})$ increases linearly with ε . As ε approaches B_{o} , thin weathering zones are favored, causing the flux to decrease rapidly with increasing erosion rate. The rapidity with which this decrease occurs depends on $k_{i}A_{i}/B_{b}$ —large values of $k_{i}A_{i}/B_{b}$ cause this transition to occur over a small range of ε , while small values of $k_{i}A_{i}/B_{b}$ result in a gradual transition to lower fluxes as ε approaches B_{o} (Fig. 3B). There is thus a maximum value of ${}_{p}F_{i}/({}_{o}q_{i}r_{i})$ that occurs at intermediate erosion rates in which the mineral supply is relatively rapid, but residence time of minerals within the weathering zone is sufficiently long to weather the available supply (Ferrier and Kirchner, 2008; Gabet and Mudd, 2009) (Fig. 3B).

Early workers (Gilbert, 1880) recognized that soil production may depend on the establishment of a layer that can retain water, and so if rapid erosion thins the soil to a great extent, a positive feedback may be established that thins the soil. In this case, Eq. (4) may not be wholly appropriate at rapid denudation rates, since soil in these circumstances may be stripped entirely from certain portions of the landscape. We speculate that such an effect will be expressed in a similar manner to that shown in Fig. 3B, although the onset of soil thinning at lower erosion rates may cause ${}_{p}F_{i}/({}_{o}q_{i}r_{i})$ to decrease more rapidly.

Finally, in the case that topographic relief sets a limit on the thickness of the weathering zone, D_i is represented by Eqs. (11) and (12). In the simple case in which relief is linearly related to ε (Eq. (12)), the extent of weathering is independent of ε , while the weathering flux produced by a particular mineral phase retains this dependence. If relief increases nonlinearly with ε , as has been shown to be the case in rapidly denuding environments (Eq. (11)), the extent of mineral weathering retains some weak inverse dependence on ε .

3.2. Erosion rates versus reaction kinetics

Thus far, we have focused exclusively on illuminating the control that the relative timescale of erosion versus reaction kinetics plays in



Fig. 2. (A) Relationship between D_i and pr_i^{ϵ} . (B) The competition between mineral supply and reaction can be visualized by contouring ${}_{p}F_i/({}_{o}q_ir_i)$ for each value of ε and k_iA_iZ . Gray shaded areas show combinations of ε and k_iA_iZ for which >95% of changes in ${}_{p}F_i/({}_{o}q_ir_i)$ were generated by changes in ε and k_iA_iZ for supply- and reaction-limited conditions, respectively.



Fig. 3. (A) Variation of ${}_{p}F_{i}^{*}$ with ε^{*} for different values of the dimensionless parameter $(k_{i}A_{i})/(B_{o}B_{b})$. (B) Influence of ε_{c} , B_{o} , and $(k_{i}A_{i})/B_{b}$ on values of ${}_{p}F_{i}/({}_{o}q_{i}r_{i})$. Red, yellow, and green lines show the effect of changing B_{o} , while solid, dashed, and dotted lines show the effect of changing $(k_{i}A_{i})/B_{b}$.

controlling silicate-weathering rates. However, field- and laboratorybased studies provide ranges of the timescales for the reaction kinetics of a variety of silicate minerals, including albite, biotite, hornblende, and potassium feldspar (compiled in White and Brantley, 1995). In addition, we have compiled those few observations for which simultaneous measurements of erosion rates (determined via cosmogenic radionuclide analyses) and weathering zone thicknesses exist to provide plausible bounds on the values of B_o and B_b in Eq. (4) (Hilley and Porder, 2008). Importantly, while there are large uncertainties in the use of Eq. (4), we used a range that bounded all available data, and estimates of these parameters encapsulate a large range of the parameter space of ε and Z. Thus, while specific values of these parameters are likely incorrect in detail, the bounds used as part of our previous work (Hilley and Porder, 2008) explain all data currently available. Using these two timescales, we can estimate the degree to which the weathering flux of each mineral phase is dominated by supply- or reaction-limited conditions by determining the ratio of the chemical weathering flux of each mineral



Fig. 4. Measured (dots) and modeled (solid and dashed lines) weathering zone thickness as a function of erosion rate. Lines labeled "Exponential Weathering Zone Model" assume the functional form of Eq. (4) and use field observations to calibrate the empirical constants B_o and B_b . "Relief Limit to Weathering Zone" shows the case in which the vertical extent of the weathering zone is limited by the amount of relief, as might be expected for low-relief landscapes in which weathering zone thickness is controlled primarily by the water balance. For the case that relief limits the extent of the weathering zone, we consider only measurements taken from erosion rates >0.01 mm/yr to bound the relationship. Thus, the upper bound of modeled weathering zone thicknesses for this scenario is shown as the dashed line.

phase relative to its total mass flux ($_pF_i^*$), and determining how this changes with erosion rate.

We consider two weathering zone thickness models corresponding to the different ways we conceive weathering zone thickness to be related to erosion rate, as described above (Fig. 4). In the first model, we use data presented in Hilley and Porder (2008) to calibrate the empirical constants in Eq. (4) that bound all compiled observations, and refer to this scenario as "Exponential Weathering Zone Model". For the second case, we regarded the landscape relief as the factor that limits the vertical extent of weathering in the subsurface. We used the relationship between erosion rate and relief presented in Montgomery and Brandon (2002) and Hilley and Porder (2008) to calculate weathering zone thickness as a function of erosion rate. However, the large relief observed at high erosion rates produces weathering zone thicknesses that are incompatible with field measurements. To account for both the limiting effect of relief on the extent of the weathering zone as well as field-based observations, we calibrated an exponential weathering zone model using observations from areas in which erosion rates were >0.01 mm/yr, and limited the maximum vertical extent of the weathering zone using the relationship between local relief and erosion rate (Fig. 4). This second scenario is referred to as "Exponential Weathering Zone Model Limited By Relief" (Fig. 4).

Finally, to gain a crude measure of the distribution of worldwide erosion rates, we used the SRTM-90 meter dataset to calculate local relief in an identical way as Montgomery and Brandon (Montgomery and Brandon, 2002; Hilley and Porder, 2008). This approach necessarily overestimates erosion rates in low-relief, depositional areas where our model does not apply. Mean erosion rates predicted by our analysis are ~0.017 mm/yr, with 90% of erosion rates falling between ~0.0017 mm/yr and ~0.15 mm/yr.

Using these estimated erosion rates and weathering zone thickness scenarios, we estimated $_{p}F_{i}^{*}$ for the range of reported kinetic



Fig. 5. Relationship between erosion rate and fraction of weathering flux controlled by mineral supply for (A) field- and (B) laboratory-based kinetic reaction rates, considering the exponential weathering zone model described in text. This relationship changes when (C) field- and (D) laboratory-based kinetic reaction rates are used with a weathering zone whose extent is limited by landscape relief at low erosion rates. Different colored envelopes represent different mineral phases. The range encapsulated in these envelopes represents the variation observed in both kinetic rate constants and inferred empirical constants relating weathering zone thickness to erosion rates. For reference, the cumulative probability distribution of erosion rates is shown as bold solid line and can be used to infer the percentage of Earth's surface subject to supply ($_{p}F_{i}^{*}>0.5$) or reaction ($_{p}F_{i}^{*}<0.5$) limited conditions. An example of this is shown for the upper bound of Kspar, where we expect the threshold between reaction- and supply-limited conditions to occur at a denudation rate of 0.05 mm/yr. Approximately 71% of the land surface is eroding at this rate or slower, which will produce conditions that may be influenced more by supply of fresh minerals than reaction kinetics.

rate constants inferred from field studies (Fig. 5A, C) and laboratory studies (Fig. 5B, D). When $_{n}F_{i}^{*}=0.5$, mineral supply and reaction kinetics play equal roles in moderating the silicate-weathering flux. The envelopes in Fig. 5 reflect the ranges in kinetic constants reported in White and Brantley (1995) and the inferred range of B_0 , B_b , and k_1 for the scenarios shown in Fig. 4. We see that in the case of laboratory kinetic rate constants, the rapid timescale of dissolution relative to residence time within the weathering zone causes all of these mineral phases to be strongly supply-limited over a broad range of denudation rates (Fig. 5B, D). When using field-based kinetic rate constants with an exponential weathering zone model unlimited by relief, high values of denudation rates (corresponding to low residence times) produced reaction-limited conditions (Fig. 5A). Thus, we expect those areas undergoing slow denudation $(<1 \times 10^{-4} \text{ mm/yr})$ to be most sensitive to changes in denudation rate, while rapidly denuding areas (>0.5 mm/yr) will be sensitive to changes in factors that alter the reaction kinetic rate constants. If relief limits the extent of the weathering zone at low denudation rates, weathering fluxes depend on the individual mineral phase of interest, but generally tend to be more reaction-limited than expected using the unrestricted weathering zone model. Despite the weathering zone model employed, at low erosion rates, weathering fluxes depend on erosion rate to some degree, and depending on the kinetics that are applicable to a specific area, weathering may depend as much on fresh mineral supply as reaction kinetics. In addition, at high denudation rates, residence times within the weathering zone are expected to be short, and because fresh minerals weather more rapidly than do those subjected to lengthy weathering (White and Brantley, 1995), chemical weathering may be slightly more rapid than predicted by our simple model under these circumstances. In addition, enhancement of the kinetic reaction rates by plant-derived soil acidification may cause weathering rates to be more like those observed in the laboratory. Conversely, the effects of pore-water saturation may systematically slow reaction rates over time, causing them to be better represented by field-based reaction rates.

In Fig. 5, we show the cumulative distribution function of erosion rates, and plot the fraction of the Earth experiencing erosion rates less than or equal to each value of erosion rate as a solid dark line. This allows us to estimate the fraction of Earth's surface over which erosion or kinetic reaction rates may be more or less important. Based on these estimates of global erosion rates, we note the percentage of the world in which $_{\nu}F_{i}^{*}$ > 0.5 in Table 2, indicating a weakly to strongly dominant influence of mineral supply on weathered fluxes of each mineral phase investigated. We find that given the state-of-knowledge of kinetic reaction rates, denudation rates, and the relationship between weathering zone thickness and denudation rates, denudation may play an important role in moderating silicate-weathering fluxes for large portions of the world for the four silicate mineral phases investigated. These results suggest that reaction kinetics plays an important role in moderating silicate-weathering fluxes; however, fresh minerals supplied by tectonic and erosional processes may play a sub-equal or greater role in moderating silicate-weathering rates under some circumstances.

Table	2

Pe	ercentage	of	weathering	attri	butab	le	to	effects	of	mineral	supp	oly
----	-----------	----	------------	-------	-------	----	----	---------	----	---------	------	-----

Mineral phase	Field-based rate	Field-based rate	Lab-based rate		
	Exponential Z	Exponential Z	Both weathering		
	Model	Limited by relief	Zone scenarios		
Albite K-feldspar Hornblende Biotite	48-100% 6-71% 9-77% 54-100%	52–100% 0–71% 0–77% 57–100%	100% 99–100% 100% 100%		

4. Discussion

The idea that mineral supply might regulate mineral weathering rates has a rich history in the weathering literature. Initially, Stallard and Edmond (1983) observed that denudation plays an important role in moderating chemical fluxes from the Amazon Basin. This observation partly led to the hypothesis that the uplift of large mountain belts such as the Himalayas may accelerate chemical weathering processes by creating reactive surface area, which in turn may alter the composition of Earth's atmosphere as the release of Ca^{2+} and Mg^{2+} sequesters CO₂ into geologic deposits (Raymo and Ruddiman, 1992; Ruddiman et al., 1997). Since these regional and global inferences have been made, cosmogenic isotope measurements have been used with geochemical analyses to relate denudation of landscape elements to the chemical weathering they undergo. To this end, a number of ground-breaking studies have shown strong covariation between denudation of Earth's surface and the chemical mass flux produced by weathering (Riebe et al., 2001b, 2003, 2004a,b), and such studies continue to appear in the literature (Green et al., 2006; Yoo et al., 2007; Burke et al., 2009; Dixon et al., 2009; Burke et al., in revision; Yoo et al., 2009). Motivated in part by these results, Waldbauer and Chamberlain (2005) developed a simple mathematical model, utilized herein, to relate the relative timescales of soil residence and reaction to determine the extent to which chemical weathering rates may be controlled by reaction kinetics versus mineral supply. Field calibration of this model at the catchment scale showed that such an approach might be used to explain regional-scale covariation between denudation and chemical weathering rates (Hren et al., 2007b). Based on these results, our previous work coupled a model of the development of the weathering zone that logarithmically related weathering zone thickness to denudation rate, and used the regionally calibrated model with global estimates of erosion rate to infer silicate-weathering fluxes at the global scale (Hilley and Porder, 2008). At the same time, Ferrier and Kirchner (2008) used a similar approach to understand how temporal fluctuations in denudation rates may alter chemical weathering fluxes. Additionally, Gabet and Mudd (2009) present a steady-state version of this model that is virtually identical to that of Waldbauer and Chamberlain (2005) and Hilley and Porder (2008). Many of these prior contributions harkened the thickness of mobile soil equivalent to that of the entire weathering zone (Ferrier and Kirchner, 2008; Gabet and Mudd, 2009), which in many natural situations, penetrates much deeper into the undisturbed rock (Hilley and Porder, 2008). Importantly, all of these studies require the weathering zone thickness to be logarithmically related to the denudation rate, and do not consider a case where the extent of the groundwater system may restrict the total thickness of the weathering zone as we have done in this study. Additionally, none explore the dynamics of individual mineral phases in the weathering process when considering our best guess as to how deep, on average, weathering might extend, as well as how fast, on average, landscapes might be denuding. In this regard, the current contribution clarifies the relative role that mineral supply and reaction may play for a variety of different mineral phases through dimensional analysis and the inclusion (albeit in a coarse way) of the limiting effects of groundwater flow on the extent of the weathering zone in the subsurface.

Several studies have argued that the uplift of mountain belts may facilitate accelerated chemical weathering rates by increasing the supply of minerals available to be weathered (Raymo et al., 1988; Raymo and Ruddiman, 1992; Ruddiman et al., 1997). If the supply of fresh minerals by erosional lowering of topography is as large a contributor to overall weathering rates as our analysis suggests, active mountain belts may indeed contribute a disproportionate fraction to the silicate-weathering budget (Stallard and Edmond, 1983; Hren et al., 2007a; Hilley and Porder, 2008). Furthermore, in areas of exceedingly rapid denudation, residence times of rock within the weathering zone may be low enough to cause a decrease in the weathering fluxes (Fig. 3), although rapid weathering of freshly exposed rock such as is found in landslide scars may augment weathering of silicate minerals in these circumstances (Gabet, 2007). We stress that a key, and poorly constrained factor in our analysis is the relationship between denudation rate and weathering zone thickness, and so more data that elucidate how weathering zone thickness varies in different climatic and tectonic conditions are required to form strong conclusions regarding the strength of coupling between erosion rates and chemical weathering. Nonetheless, based on our current state-of-knowledge, it appears that accelerated erosion in the world's active orogens is capable of driving accelerated silicate weathering, consistent with the idea that the uplift of the Himalayas and Tibet has the potential to change Earth's atmosphere and climate during the Cenozoic (Raymo et al., 1988; Raymo and Ruddiman, 1992; Ruddiman et al., 1997).

5. Conclusions

This contribution uses a simple mathematical model of the kinetics of mineral weathering and the effects of fresh mineral supply by surface denudation to examine the potential role of each in controlling chemical weathering rates. Examination of the model reveals that when the timescale of weathering reactions (defined by mineral specific kinetics, $(k_iA_i)^{-1}$) is much shorter than the time during which minerals weather in the near-surface (defined by weathering zone thickness and denudation rate, Z/ε), primary minerals will be exhausted and mineral supply will limit weathering rates. Conversely, if the timescale of mineral weathering reactions is much longer than the time during which minerals are allowed to weather, the reaction kinetics will instead control weathering rates. Additionally, our analysis highlights that the weathering time of minerals near the surface depends on both weathering zone thickness (Z) and denudation rate (ε). We examined several plausible scenarios that relate weathering zone thickness to denudation rate. In the first, the thickening rate of the weathering zone is regarded analogous to that of soils, with thinner weathering zones associated with more rapid denudation. In this case, chemical weathering rates reach a maximum value at intermediate denudation rates, which produce relatively thick weathering zones while maintaining rapid mineral supply rates. In the second scenario, weathering zone thickness is limited by the extent of the vadose zone, which we assume is proportional to the amount of local topographic relief. In this case, weathering rates increase approximately linearly with denudation rates as the effect of changing weathering zone thickness is offset by denudation rate to maintain a relatively constant fraction of primary minerals weathered.

We combined the range of $k_i A_i$ measured in the laboratory or inferred from field studies for four silicate minerals with a range of plausible denudation rates and weathering zone thicknesses to determine those circumstances that might produce chemical weathering rates limited by mineral supply versus reaction kinetics. Rapid mineral weathering kinetics observed in the laboratory produce supply-limited conditions over a large spectrum of plausible erosion rates, while field-based kinetics shows that mineral supply may limit weathering rates in cases of slow erosion. We stress that a key, and poorly constrained factor in our analysis is the relationship between denudation rate and weathering zone thickness, and so more data that shed light on how weathering zone thickness varies in different climatic and tectonic environments are required to reach strong conclusions about the relative role of reaction kinetics versus mineral supply limitation in natural systems. However, based on those data that exist, our analysis suggests that increasing denudation rates may be capable of driving accelerated silicate weathering, consistent with the idea that uplift of the Himalayas and Tibet has the potential to change Earth's atmosphere and climate during the Cenozoic (Raymo et al., 1988; Raymo and Ruddiman, 1992; Ruddiman et al., 1997).

Appendix A. Model parameters used in text

Parameter	Description	Units			
Ζ	Thickness of weathering zone	[L]			
3	Denudation rate	$[Lt^{-1}]$			
k _i	Kinetic rate constant	$[mol L^{-2} t^{-1}]$			
Ai	Mineral surface area	[L ² mol]			
oqi	Concentration of <i>i</i> th phase in bedrock	$[mol L^{-3}]$			
q_i	Concentration of the <i>i</i> th phase	$[mol L^{-3}]$			
ri	Stoichiometric ratio				
$_{p}F_{i}$	Product flux of <i>i</i> th phase	$[mol L^{-2} t^{-1}]$			
$B_{o_{\text{soil}}}$	Surface soil production rate	$[Lt^{-1}]$			
B _{b_{soil}}	Depth scaling of soil production	$[L^{-1}]$			
R _f	Local relief	[L]			
<i>k</i> ₁	Relief-denudation rate scaling factor	$[L^{(1-k_2)}t^{-1}]$			
k ₂	Relief-denudation rate exponent				
Dimensionless groups:					
$_{n}F_{i}^{*}$	Dimensionless product flux				
E.	Ratio of denudation to surface				
	Weathering zone lowering rate				
Di	Erosional Damköhler number				

References

- Ahnert, F., 1970. Functional relationship between denudation, relief, and uplift in large mid-latitude drainage basins. Am. J. Sci. 268, 243–263.
- Berner, R., 1990. Atmospheric carbon dioxide levels over Phanerozoic time. Science 249, 1382.
- Berner, R.A., 1994. 3GEOCARBIII: a revised model of atmospheric CO₂ over Phanerozoic time. Am. J. Sci. 294, 56–91.
- Berner, R.A., Lasaga, A.C., Garrels, R.M., 1983. The carbonate-silicate geochemical cycle and its effect on atmospheric carbon dioxide over the past 100 million years. Am. J. Sci. 283, 641–683.
- Brady, P.V., Carroll, S.A., 1994. Direct effects of CO₂ and T on silicate weathering: possible implications for climate control. Geochim. Cosmochim. Acta 58, 1853–1856.
- Brantley, S.L., 2008. Understanding soil time. Science 321, 1454-1455.
- Brantley, S.L., Crane, S.R., Creear, D., Hellmann, R., Stallard, R., 1986. Dissolution at dislocation etch pits in quartz. Geochim. Cosmochim. Acta 50, 2349–2361.
- Brantley, S.L., Blai, A.C., Cremeens, D.L., MacInnis, I., Darmody, R.G., 1993. Natural etching rates of feldspar and hornblende. Aquat. Sci. 55, 262–272.
- Burke, B., Heimsath, A.M., Chappell, J., Yoo, K., 2009. Weathering the escarpment: chemical and physical rates and processes, southeastern Australia. Earth Surf. Processes Landf. doi:10.1002/esp. 1764.
- Burke, B., Heimsath, A. M., Kaste, J. M., in revision. Quantifying the spatial variability of chemical weathering across soil-mantled landscapes. Geoderma.
- Carson, M.A., Kirkby, M.J., 1972. Hillslope Form and Process. Cambridge University Press, London.
- Chamberlain, C.P., Waldbauer, J.R., Jacobson, A.D., 2005. Strontium, hydrothermal systems and steady-state chemical weathering in active mountain belts. Earth Planet. Sci. Lett. 238, 351–366.
- Clow, D.W., Drever, J.I., 1996. Weathering rates as a function of flow through an alpine soil. Chem. Geol. 132, 131–141.
- Dietrich, W.E., Wilson, C.J., Montgomery, D.R., McKean, J., Bauer, R., 1992. Erosion thresholds and land surface morphology. Geology 20, 675–679.
- Dietrich, W.E., Bellugi, D.G., Sklar, L.S., Stock, J.D., Heimsath, A.M., Roering, J.J., 2003. Geomorphic transport laws for predicting landscape form and dynamics. In: Wilcock, P.R., Iverson, R.M. (Eds.), Prediction in Geomorphology: Geophysical Monograph, vol. 135, pp. 103–132.
- Dixon, J.L., Heimsath, A.M., Kaste, J.M., Amundson, R., 2009. Climate driven processes of hillslope weathering. Geology.
- Drever, J.I., Clow, D.W., 1995. Weathering rates in catchments. Rev. Mineral. 31, 463–483.
- Evans, M.J., Derry, L.A., Anderson, S.P., France-Lanord, C., 2001. Hydrothermal source of radiogenic Sr to Himalayan rivers. Geology 29, 803–806.
- Ferrier, K.L., Kirchner, J.W., 2008. Effects of physical erosion on chemical denudation rates: a numerical modeling study of soil-mantled hillslopes. Earth Planet. Sci. Lett. 272, 591–599.
- Gabet, E.J., 2007. A theoretical model coupling chemical weathering and physical erosion in landslide-dominated landscapes. Earth Planet. Sci. Lett. 264, 259–265.
- Gabet, E.J., Mudd, S.M., 2009. A theoretical model coupling chemical weathering rates with denudation rates. Geology 37, 151–154.
- Gilbert, G.K., 1880. Report on the Geology of the Henry Mountains. Government Printing Office, Washington, D.C.
- Green, E.G., Dietrich, W.E., Banfield, J.F., 2006. Quantification of chemical weathering rates across an actively eroding hillslope. Earth Planet. Sci. Lett. 242, 155–169.
- Heimsath, A.M., Dietrich, W.E., Nishiizumi, K., Finkel, R.C., 1997. The soil production function and landscape equilibrium. Nature 388 (6640), 358–361.
- Heimsath, A.M., Dietrich, W.E., Nishiizumi, K., Finkel, R.C., 1999. Cosmogenic nuclides, topography, and the spatial variation of soil depth. Geomorphology 27, 151–172.

- Hilley, G.E., Porder, S., 2008. A framework for predicting global silicate weathering and CO₂ drawdown rates over geologic time-scales. Proc. Natl. Acad. Sci. 105, 16855–16859.
- Hren, M.T., Chamberlain, C.P., Hilley, G.E., Blisniuk, P.M., Bookhagen, B., 2007a. Major ion chemistry of the Yarlung Tsangpo–Brahmaputra river: chemical weathering, erosion, and CO₂ consumption in the southern Tibetan plateau and eastern syntaxis of the Himalaya. Geochim. Cosmochim. Acta 71, 2907–2935.
- Hren, M.T., Hilley, G.E., Chamberlain, C.P., 2007b. The relationship between tectonic uplift and chemical weathering rates in the Washington Cascades: field measurements and model predictions. Am. J. Sci. 307, 1041–1063.
- Lasaga, A.C., 1998. Kinetic Theory in the Earth Sciences. Princeton University Press, Princeton, N. J.
- Lasaga, A.C., Soler, J.M., Ganor, J., Burch, T.E., Nagy, K.L., 1994. Chemical weathering rate laws and global geochemical cycles. Geochim. Cosmochim. Acta 58 (10), 2361–2386.
- Lebedeva, M.I., Fletcher, R.C., Balashov, V.N., Brantley, S.L., 2007. A reactive diffusion model describing transformation of bedrock to saprolite. Chem. Geol. 244, 624–645.
- Lebedeva, M., Fletcher, R., Brantley, S., in press. A mathematical model for steady-state regolith production at constant erosion rate. Earth Surface Processes and Landforms.
- Maher, K., Steefel, C.I., Stonestrom, D.A., 2007. Control of chemical weathering rates by secondary mineral precipitation: a reactive transport approach for understanding soil genesis. Geochim. Cosmochim. Acta 71, A613.
- Montgomery, D.R., Brandon, M.T., 2002. Topographic controls on erosion rates in tectonically active mountain ranges. Earth Planet. Sci. Lett. 201, 481–489.
- Montgomery, D.R., Dietrich, W.E., 1988. Where do channels begin? Nature 336, 232–234.
- Montgomery, D.R., Dietrich, W.E., 1992. Channel initiation and the problem of landscape scale. Science 255, 826–830.
- Porder, S., Hilley, G.E., Chadwick, O.A., 2007. Chemical weathering, mass loss, and dust inputs across a climate by time matrix in the Hawaiian Islands. Earth Planet. Sci. Lett. 258, 414–427.
- Raymo, M.E., Ruddiman, W.F., 1992. Tectonic forcing of late Cenozoic climate. Nature 359, 117–122.
- Raymo, M.E., Ruddiman, W.F., Froelich, P.N., 1988. Influence of late Cenozoic mountain building on ocean geochemical cycles. Geology 16, 649–653.
- Riebe, C.S., Kirchner, J.W., Granger, D.E., Finkel, R.C., 2001a. Minimal climatic control on erosion rates in the Sierra Nevada, California. Geology 29, 447–450.
- Riebe, C.S., Kirchner, J.W., Granger, D.E., Finkel, R.C., 2001b. Strong tectonic and weak climatic control of long-term chemical weathering rates. Geology 26, 511–514.
- Riebe, C.S., Kirchner, J.W., Finkel, R.C., 2003. Long-term rates of chemical weathering and physical erosion from cosmogenic nuclides and geochemical mass balance. Geochim. Cosmochim. Acta 67, 4411–4427.

- Riebe, C.S., Kirchner, J.W., Finkel, R.C., 2004a. Erosional and climatic effects on longterm chemical weathering rates in granitic landscapes spanning diverse climate regimes. Earth Planet. Sci. Lett. 224, 547–562.
- Riebe, C.S., Kirchner, J.W., Finkel, R.C., 2004b. Sharp decrease in long-term chemical weathering rates along an altitudinal transect. Earth Planet. Sci. Lett. 218, 421–434.
- Ruddiman, W.F., Raymo, M.E., Prell, W.L., Kutzbach, J.E., 1997. The uplift-climate connection: a synthesis. In: Ruddiman, W.F. (Ed.), Tectonic Uplift and Climate Change. Plenum Press, New York, pp. 471–515.
- Stallard, R.F., Edmond, J.M., 1983. Geochemistry of the Amazon 2. The influence of geology and weathering environment on the dissolved load. J. Geophys. Res. 88, 9671–9688.
- Turcotte, D.L., Schubert, G., 2002. Geodynamics, 2nd Edition. Cambridge University Press, New York.
- Velbel, M.A., 1993. Temperature dependence of silicate weathering in nature: how strong a feedback on long-term accumulation of atmospheric CO₂ and global green house warming? Geology 21, 1059–1062.
- Waldbauer, J.R., Chamberlain, C.P., 2005. Influence of uplift, weathering and base cation supply on past and future co₂ levels. In: Ehleringer, J.R., Cerling, T.E., Dearing, M.D. (Eds.), A History of Atmospheric CO₂ and Its Effects on Plants, Animals, and Ecosystems. : Ecological Studies, Vol. 177. Springer Science, New York, pp. 166–184.
- West, A.J., Galy, A., Bickle, M., 2005. Tectonic and climatic controls on silicate weathering. Earth Planet. Sci. Lett. 235, 211–228.
- White, A.F., 1995. Chemical weathering rates of silicate minerals in soils. Rev. Mineral. 31, 409–461.
- White, A.F., Blum, A.E., 1995. Effects of climate on chemical weathering rates in watersheds. Geochim. Cosmochim. Acta 59, 1729–1747.
- White, A.F., Brantley, S.L., 1995. Chemical weathering rates of silicate minerals: an overview. Rev. Mineral. 31, 1–22.
- White, A.F., Brantley, S.L., 2003. The effect of time on the weathering of silicate minerals: why do weathering rates differ in the laboratory and field? Chem. Geol. 202, 479–506.
- White, A.F., Blum, A.E., Schultz, M.S., Bullen, T.D., Harden, J.W., Peterson, M.L., 1996. Chemical weathering rates of a soil chronosequence on granitic alluvium: I. Quantification of mineralogic and surface area changes and calculation of primary silicate reaction rates. Geochim. Cosmochim. Acta 60, 2533–2550.
- White, A.F., Bullen, T.D., Schultz, M.S., Blum, A.E., Huntington, T.G., Peters, N.E., 2001. Differential rates of feldspar weathering in granitic regoliths. Geochim. Cosmochim. Acta 65 (6), 847–869.
- Yoo, K., Amundson, R., Heimsath, A.M., Dietrich, W.E., Briimhall, G.H., 2007. Integration of geochemical mass balance with sediment transport to calculate rates of soil chemical weathering and transport on hillslopes. J. Geophys. Res. 112, F02013.
- Yoo, K., Mudd, S.M., Sanderman, J., Amundson, R., Blum, A., 2009. Spatial patterns and controls of soil chemical weathering rates along a transient hillslope. Earth Planet. Sci. Lett. 288, 184–193.